THE ICE STORM OF JANUARY 7-10, 1956 OVER THE NORTHEASTERN UNITED STATES

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1. INTRODUCTION

The first major storm of 1956 affecting the northern Atlantic Coastal States was an intense one of the "northeaster" type which caused an extensive area of freezing precipitation. On January 7, the weather patterns of the storm were developing both at the surface and aloft over the Atlantic Ocean and North America. A strong block south of Greenland and an intense Low east of Hatteras combined to produce an area of glaze which began as sleet and freezing rain shortly before midnight on the 7th in eastern Maine and spread rapidly westward and southward. The southern limit of these hydrometeors was reached on the 10th in the Carolinas and Tennessee with the westernmost limit occurring in Michigan on the same date. The duration at any one station was quite variable, ranging from 5 to 15 hours in the coastal States to over 75 hours (intermittently) in portions of Ohio. Only the rapid transition of temperatures from below freezing to thawing conditions kept this storm from becoming one with major ice damage. However, from onset to the very end, these freezing types of precipitation persisted for seven days over some portion of the northeastern States. In this article we shall for the most part confine our discussion to the first 4 days.

2. ANTECEDENT CONDITIONS

During the 5 weeks preceding this storm, the storm tracks over the eastern portion of the nation were primarily along the northern border of the country. Under such conditions, precipitation totals had been small and generally near or below normal temperatures had prevailed. (See Andrews [1].)

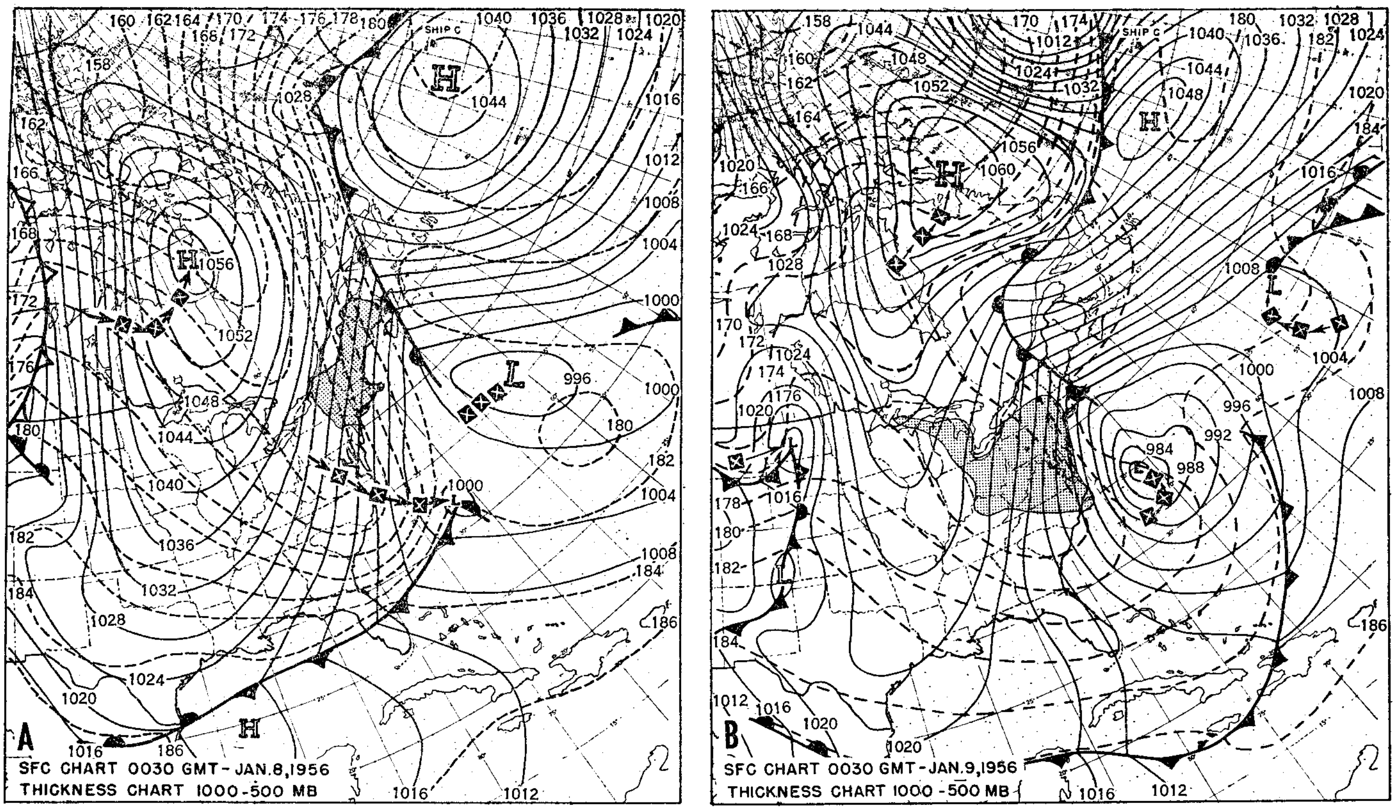
The low pressure system which was partially responsible for the ice storm had its inception along the Pacific polar front on January 4 when it first appeared in association with a short-wave trough aloft that was moving out of the long-wave position just off the west coast. Advancing along the northern border of the United States just south of the thermal field of an Arctic front, the Low remained weak as it passed through the long-wave ridge position over central United States. A new development occurred at the point of occlusion on the 6th, in the vicinity of Lake Michigan, as the original center weakened and dissipated over Lake Superior. The eastward movement continued through the 6th, and by 0030 GMT on the 7th the Low had reached the Pittsburgh, Pa. area.

Off the Atlantic shore was the decadent remains of a low pressure system which had aided in the transporting of cold air southward through the Ohio Valley to the Gulf of Mexico and eastward over the southern Atlantic Coast States. This cyclone had remained off the eastern seaboard of the United States for several days but was now beginning to move slowly eastward.

3. SYNOPTIC FEATURES, JANUARY 7 TO 10

On the morning of January 7, a surge of Arctic air was observed moving southward from Canada into the United States. With pressures building and gradients tightening to the west and north of the cyclone located near Pittsburgh, much of the continent from north of Hudson Bay southward to the Gulf of Mexico was under the domination of a large anticyclone centered just south of Churchill, Manitoba. In the upper air a building ridge extended north-northeastward from Texas to Hudson Bay. As the dominant surface anticyclone and ridge aloft turned the winds to a more northerly direction, the Pittsburgh Low, as well as the associated cut-off Low aloft near Buffalo, was now changing its course to a more southeasterly direction. At the same time a well-developed blocking High pressure cell was centered near stationary ship "C" (52°45′ N., 35°30′ W.), but its retrogression toward Newfoundland was clearly indicated by intense deepening in the Iceland to British Isles area.

The strong confluent northwestward flow of marine air between the pressure systems in the Atlantic had carried the warm front to the north of the now-dying cyclone. Nevertheless, the front remained well-defined, with the temperature differential across it very sharp both at the surface and aloft. The maritime provinces of Nova Scotia and Newfoundland reported temperatures in the upper 40's while readings on the continent were in the 10° F. to 20° F. range. Aloft the 1000–500-mb. thickness for 0300 GMT of the 7th indicated a strong thermal gradient west of this front, approximately 700 feet in 180 miles. This gradient would indicate thermal winds of about 100 knots, in close agreement with the 110-knot thermal wind computed from the Quebec rawin.



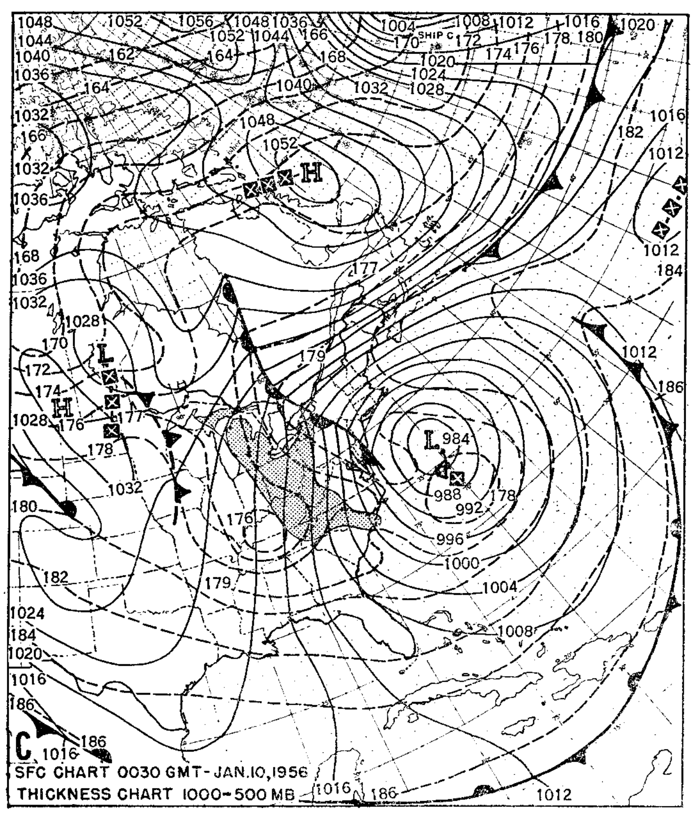


FIGURE 1.—Sea level and 1000–500-mb. thickness synoptic patterns for January 8–10, 1956. Shaded area represents region in which freezing rain or sleet occurred in the 24-hour period 0001 to 2400 EST of each date. (A) 0030 GMT surface chart for January 8, 1956 with 1000–500-mb. thickness for 0300 GMT in dashed lines at 200-foot intervals. (B) 0030 GMT surface chart for January 9, 1956 with 1000–500-mb. thickness for 0300 GMT. (C) 0030 GMT surface chart for January 10, 1956 with 1000–500-mb. thickness for 0300 GMT.

By 0030 GMT of January 8, the Pittsburgh Low had crossed the coast in the vicinity of Cape Hatteras and was located at a point 300 nautical miles east-southeast of that area (fig. 1A). The 500-mb. low center of 17,500 feet was over the Hatteras region (fig. 2A). Simultaneously, the blocking conditions south of Greenland had begun to retrograde. The relative movement of these two systems served to produce a general tightening of the pressure gradient along the New England and Nova Scotia coasts. Thus, with the increased easterly wind flow the stationary front off the coast began moving westward as a warm front. Shortly after the 0030 GMT synoptic chart freezing rain and sleet began over the southeastern half of Maine. Over the interior of Canada the surface high pressure center moved into the eastern portion of Hudson Bay and continued to intensify. And at the same time the 500-mb. ridge built northward with the 18,200-foot

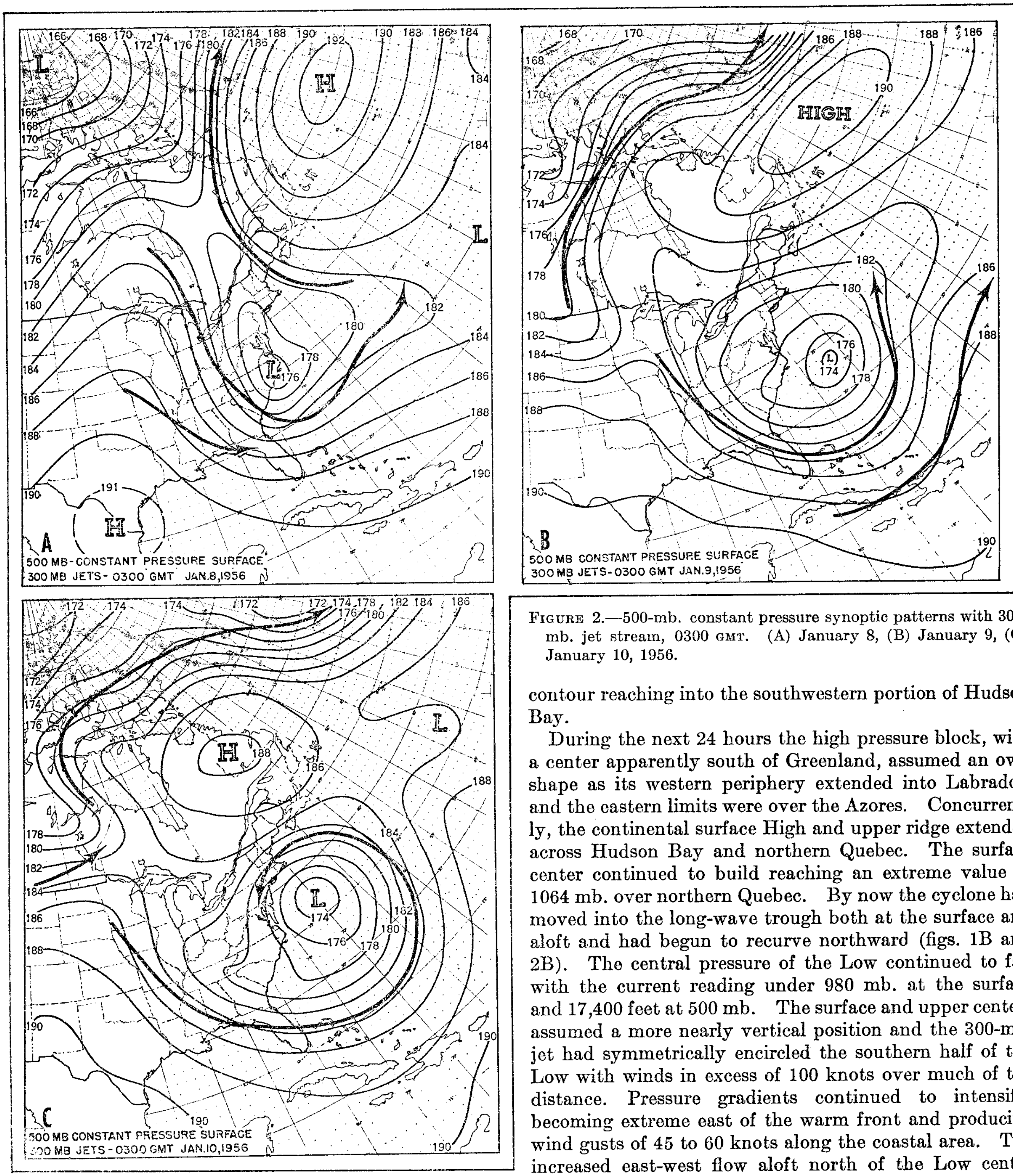


FIGURE 2.—500-mb. constant pressure synoptic patterns with 300-mb. jet stream, 0300 GMT. (A) January 8, (B) January 9, (C) January 10, 1956.

contour reaching into the southwestern portion of Hudson Bay.

During the next 24 hours the high pressure block, with a center apparently south of Greenland, assumed an oval shape as its western periphery extended into Labrador, and the eastern limits were over the Azores. Concurrently, the continental surface High and upper ridge extended across Hudson Bay and northern Quebec. The surface center continued to build reaching an extreme value of 1064 mb. over northern Quebec. By now the cyclone had moved into the long-wave trough both at the surface and aloft and had begun to recurve northward (figs. 1B and 2B). The central pressure of the Low continued to fall with the current reading under 980 mb. at the surface and 17,400 feet at 500 mb. The surface and upper centers assumed a more nearly vertical position and the 300-mb. jet had symmetrically encircled the southern half of the Low with winds in excess of 100 knots over much of the distance. Pressure gradients continued to intensify, becoming extreme east of the warm front and producing wind gusts of 45 to 60 knots along the coastal area. The increased east-west flow aloft north of the Low center carried the ice shield farther inland. It had, by midnight EST of the 8th, covered part of New Jersey, extreme northeastern Pennsylvania, and more than half of New York State. (See shaded area on fig. 1B.)

The warm front had continued moving westward and was by then crossing Massachusetts and New Hampshire (fig. 3). Following the passage of the warm front, temperatures rose rapidly, sleet and freezing rain ended, but rain continued. The rapid transition from freezing to thawing conditions, even during night hours, was well illustrated at Danbury, Conn., where at 6 p. m. EST of the 8th the temperature was 15° F. and by 9:30 a. m. of the 9th it was 43° F. The duration of the icing conditions over the coastal States varied generally from 5 to 15 hours. The post-frontal rainfall, it is thought, was in part due to a weak tropical warm front that was overlying the area. Further discussion on this subject will appear later in this article.

By 0030 GMT of the 10th the warm front had reached a position extending from Delaware Bay northwestward across western New York and on into Canada (fig. 3). Sleet and freezing rain continued to spread westward and southward in the strong easterly flow, the limits reached by midnight EST of the 9th extending into northern North Carolina, northeastern Kentucky, western Ohio, and Michigan. Light to locally heavy amounts of rain continued to fall east of the warm front as temperatures remained in the upper 30's or in the 40's. The synoptic situation responsible for these conditions had changed but slightly during the past 24 hours (fig. 1C) with the cold core Low having become vertical and with little change in intensity as it slowly migrated northward. At the same time the northern high pressures combined at both surface and aloft, extending the block to the Labrador region, with a ridge persisting southeastward toward the Azores. The 300-mb. jet (fig. 2C) by now had completely encircled the Low and was coming inland over eastern Massachusetts and on into Pennsylvania.

By 0030 GMT of the 11th, the pressure systems were deteriorating although blocking conditions continued. Advection of warm air both from the east and the west over the Great Lakes region (figs. 1B and C) decreased the thermal gradients finally to a point of frontolysis of the warm front. Nevertheless, overrunning continued to produce freezing rain or sleet for the next 2 days in Ohio. Sleet and freezing rain were reported in some sections of the northeastern States into the 16th, but with decreasing area. In general, the pattern was reversed after the 11th, with cooler drier air moving in from the northwest and gradually forcing the sleet and glaze conditions back across the New England States.

Much of this sleet and freezing rain occurred in a portion of the region where these types of hydrometeors are most frequent. Furthermore, as mentioned by Brooks [2], most of the ice storms in the Northeast fall into three

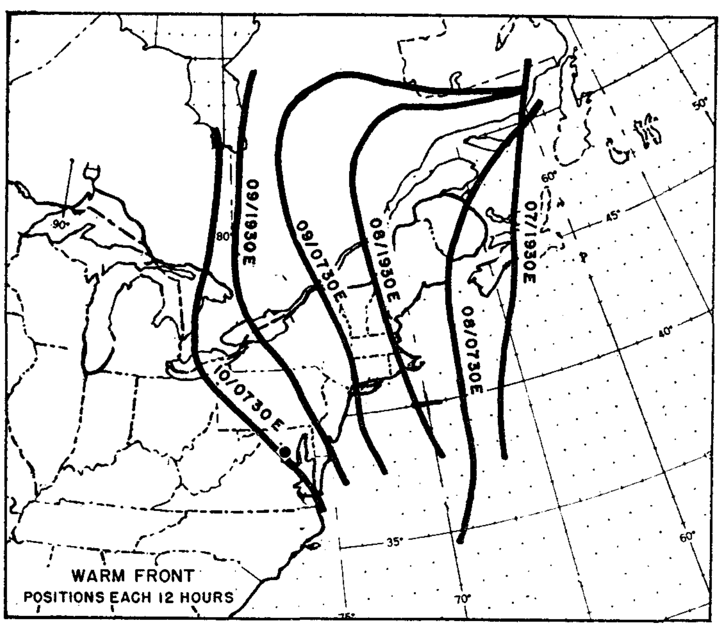


FIGURE 3.—Positions of the warm front at 12-hour intervals during the period 0030 GMT, January 8 to 1230 GMT, January 10, 1956. Times on fronts are in EST for ease in comparison with precipitation. The unusual westward movement of the warm front is clearly depicted on this chart.

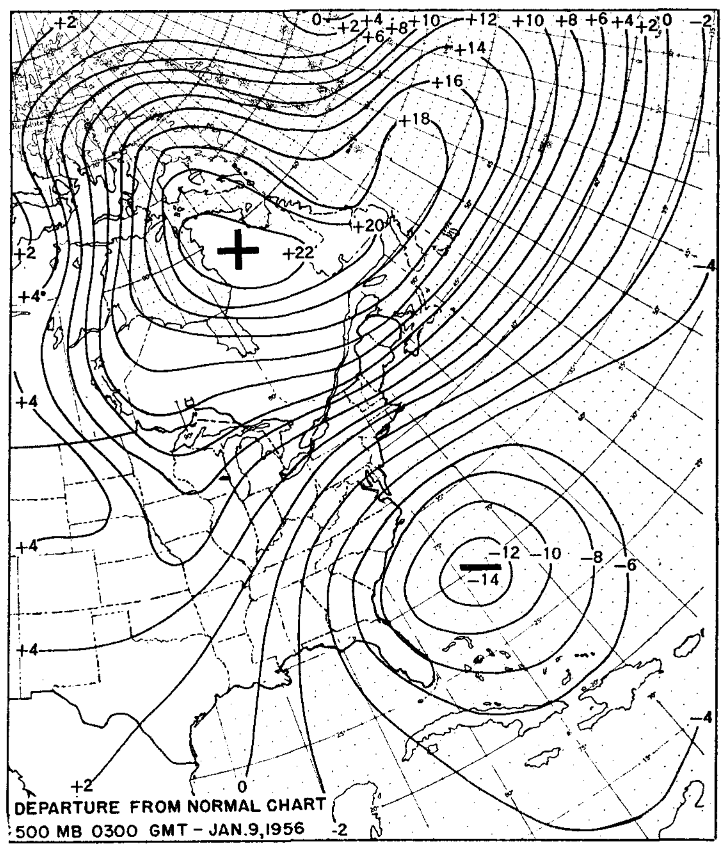


FIGURE 4.—Strength of blocking action in the Atlantic is indicated by the departures of the 500-mb. heights from normal for 0300 GMT, January 9, 1956. Values are in hundreds of geopotential feet.

classifications: (1) warm air arriving over residual cold air, (2) cold air coming in below and warm air arriving above, and (3) cold air pushing in from the north or west below a raincloud. During the period of this ice storm type 1 prevailed through January 11, and then changed to type 3 as the precipitation moved eastward again. The third type is in reality the reversal of the first two, i. e., the forms of precipitation change from rain or freezing rain to sleet and then into snow as the wedge of cold air moves into the area beneath the overcast.

4. BLOCKING ACTION OVER THE NORTHWEST ATLANTIC OCEAN

One of the most important broad-scale features contributing to the prolonged and widespread areas of sleet, freezing rain, and rain was the blocking action of the 500-mb. High in the northwestern Atlantic Ocean. The outstanding effect of this block was that it forced the cyclone along the Atlantic Coast to remain in the long-wave trough for a period of nearly seven days. This produced a strong easterly flow of moist air with a duration period sufficient to permit it to extend into Michigan and Ohio in the west and to the Carolinas and Tennessee to the south.

The strength of this block may have been due partially to the advection of warm air into the Quebec and Labrador region. There were indications during this period that tropical air was transported aloft over the easternmost New England States and portions of eastern Canada. Godson [3] has written that temperatures of −17° C. at 500 mb. are evidence of tropical air in Canada during the winter season, while Vederman [4] delineates the northern limit of tropical air at that level along the −20° C. isotherm. Temperatures at the 500-mb. level for a few of the eastern stations are shown in table 1.

5. THICKNESS PATTERN

Further indication of the extent of the warming from this storm over the Northeast is obtained from the change in thickness of the 1000–500-mb. layer (figs. 1A–C). It is generally known that a 200-foot increase in the 1000–500-mb. thickness is equivalent to a 5.4° F. rise in the mean virtual temperature of the layer. In this case, in northern New York State thickness values increased from 17,100 feet on the 8th to 18,000 feet on the 10th, and in northern Quebec the change was even greater, being from 16,400 feet to 17,600 feet. Thus, there was a thickness change of 900 and 1,200 feet, respectively, or a rise of the mean virtual temperature of the column of over 24° and 32° F., respectively; this is in fair agreement with actual surface temperature changes occurring during this warming period.

The thickness patterns over the eastern States and Canada also furnish an excellent picture of the intensity of the warm front.[1] On the morning of the 8th, for instance (fig. 1A), a tight thermal gradient is shown ahead of the warm front, with thickness values of 17,100 to 18,100 feet between Buffalo, N. Y., and Halifax, Nova Scotia, indicating a front of strong intensity. The easterly flow suggested by the 500-mb. contours (fig. 2A) indicates rapid advection of warm air over New Brunswick, Nova Scotia, and New England. By 0030 GMT of January 9 (fig. 1B) the range of thickness values was only between 17,500 and 17,900 feet, indicating a front of only moderate intensity. Twenty-four hours later (fig. 1C) the thermal gradient was almost destroyed (weak classification), and warm air had by this time been transported into the Carolinas, Indiana, Michigan, and the Hudson Bay region.

TABLE 1.—*500-mb. temperatures (°C.) at selected stations in North America for Jan. 7–10, 1956*

Station	Date and time							
	7th		8th		9th		10th	
	03	15	03	15	03	15	03	15 GMT
Moosonee, Ontario	−32	−33	−28	−30	−18	−13	−24	−19
Nitchequon, Quebec	−29	−32	−29	−21	−18	−20	----	−27
Stephenville, Newfoundland	−17	−17	−17	−16	−23	−25	−20	−19
Sable Island	−19	−14	−16	−18	----	−27	----	−20
Caribou, Maine	−25	−23	−16	−18	−20	−22	−18	−18
Portland, Maine	−26	−24	−19	−17	−21	−20	−17	−21
Nantucket, Mass	−23	−22	−21	−17	−19	−17	−22	−21
Albany, N. Y	−28	−29	−26	−21	−19	−18	−16	−22
Maniwaki, Quebec	−29	−30	−30	−20	−18	−25	−19	−18
Hempstead, N. Y	−25	−26	−24	----	−19	−17	−20	−22
Buffalo, N. Y	−29	−32	−30	−25	−21	−21	−18	−20
Washington, D. C	−23	−30	−31	−26	−22	−18	−18	−20

6. TROPOPAUSE

An interesting comparison of the extreme temperatures in the air surrounding the cyclone and anticyclone during this period is found in the height of the tropopause. As the surface and upper low centers moved off the eastern seaboard on the 7th and 8th in the vicinity of Hatteras, a tropopause value of 425 mb. or a height of 22,000 feet was reported at that station, with a temperature of −38° C. at the tropopause level. This low tropopause height and high temperature were confirmed by comparable values at Norfolk and Greensboro. At 0300 GMT on January 10 in the Goose Bay, Labrador area the tropopause level rose to a height of 42,000 feet (170 mb.) and the temperature fell to −72° C. This height persisted through January 11 at 0300 GMT.

7. DIFFERENTIAL ANALYSIS OF THE SURFACE PRESSURE CENTERS

As the 500-mb. Low was intensifying in the southward plunge during the 7th, the surface low center was moving out ahead of the upper Low and was coming under higher 1000–500-mb. thickness values. On the early morning surface and upper air charts of the 8th, the surface center was under a 500-mb. height of approximately 100 feet

[1] In the National Weather Analysis Center the criterion now used for the determination of the intensity of fronts is the shear across the front of the 1000–500-mb. thermal winds. A difference of 25–49 knots indicates a front of weak intensity, 50–74 knots moderate, and 75 knots or greater strong intensity.

higher than the preceding day, while the thickness value was 400 feet greater. By algebraic subtraction this made a net change of −300 feet at the 1000-mb. level or an indicated deepening of 12 mb. of the sea level Low. From the 8th to the 9th as the cold core of the Low became more nearly vertical, the height lines at 500 mb. lowered by 700 feet above the surface center, and the thickness decreased by 200 feet. Thus a net change of −500 feet was indicated at 1000 mb. or a deepening of the surface Low of 20 mb. That such deepening would occur was also indicated by the 500-mb. chart (fig. 2A). The packing and sharp curvature of contours over the South Carolina coastal region indicated considerable cyclonic vorticity in that region. That this vorticity would be advected eastward was indicated by the geostrophic winds on the mean flow chart for the same time.

In a similar procedure the building of the surface high pressure over eastern Canada can be ascertained. The retrogression of the 500-mb. block to a nearly vertical position over the surface High from the 7th to the 9th inclusive produced a rise of 800 feet at the 500-mb. level over the high center. At the same time the advection of the 1000–500-mb. thickness brought about a rise of 400 feet in thickness value during the same period. Thus there was a net change in the 1000-mb. level of 400 feet or intensification of 16 mb. at mean sea level.

8. DEPARTURES FROM NORMAL

An important aspect of this storm and the blocking high pressure system, aside from its extensive pattern of freezing precipitation and its southward and westward-moving warm front, was the deviation from normal in almost all meteorological categories. At the 500-mb. level the normal January contour gradient between 32° N., 74° W. and 58° N., 68° W. is approximately 2,200 feet (westerly geostrophic winds). At 0300 GMT, January 9 (see fig. 4) the gradient between the anomaly centers was approximately 3,400 feet and by 1500 GMT of that date had increased to about 3,900 feet. In other words, the easterly flow across these 26° of latitude was almost as great as the normal westerly flow. At no time during these 4 days, January 7–10 inclusive, was the gradient between the anomaly centers less than 2,000 feet. For a broad-scale picture of the anomaly pattern, reference may be made to figures 6 and 7 in the preceding article by Klein [5].

The 1000–500-mb. thickness also underwent extreme departures from normal. January 9, 1500 GMT thickness departures (not shown) illustrated this abnormality quite clearly with values 1,400 feet above normal over Quebec, and 600 feet below normal east of Florida, or a total departure of 2,000 feet from the normal between the anomaly centers.

The anomalies of the surface pressure along the eastern seaboard of the North American Continent were likewise outstanding. It suffices to state that the central pressure of the cyclone was the lowest ever recorded in January for that region. From midnight EST of the 9th, through 1230 EST of the 10th, the pressure of the low center varied between 976 and 980 mb. Lennahan [6] in an examination of the Historical Weather Map Series (1899 through 1951) found no January pressures in the same region of less than 981 mb. Simultaneously the anticyclone produced abnormally high surface pressure over northeastern Canada. Lennahan again found no pressures above 1040 mb. for January in the Canadian area, but under this extreme high pressure block the center attained a value of 1064 mb., registering above 1060 mb. from 1200 GMT, January 8 through 1200 GMT of January 9.

R. W. James [7] has computed the frequency of cyclones and anticyclones over North America and their intensities. He found that at latitude 30° to 35° N., the mean central pressure of cyclones was 1008 mb. with a standard deviation of 6.3 mb. The central value of this current storm thus was more than five standard deviations [2] from the mean.

Similarly the mean highest pressure between the latitudes 60° to 65° N., where the highest center was located, was computed to be 1035.5 mb. for winter months, with a standard deviation of 7.5 mb. The actual value at this time was 1064 mb. in the high center. Thus this January's value was computed as nearly four times the standard deviation [2] from the mean.

Another extreme value during this time occurred in the mean westerly flow in the temperate zone extending from 35° –55° N., over the Western Hemisphere. On January 8 the zonal index of the mean flow at 700 mb. was recorded as 1.3 m. p. s., and the 5-day mean from the 7th–11th was the lowest ever recorded, 2.9 m. p. s. For a complete discussion on these anomalies in the westerly flow, see the preceding discussion by Klein [5].

9. STATIONARY ASPECT OF LONG-WAVE TROUGH

By the use of the Hovmöller time-height diagram [8], the positions and persistency of the long-wave troughs, as well as some value of intensity, can be followed. The National Weather Analysis Center prepares this chart every 12 hours for latitudes of 35° and 50° N., extending westward from 10° W. to 140° E. From 1500 GMT, January 7 until 1500 GMT, January 10, the position of the long-wave trough off the Atlantic coast oscillated between 68° and 76° W., while off the Pacific Coast, the long-wave position ranged between 134° and 138° W. The wave length between these east and west coast long-wave troughs varied from 58° to 68° during the period.

The actual stationary wave length at 0300 GMT of the 7th was found to be 53°, a difference from observed of 14°. By 1500 GMT it was 65°, then lowered to 59° through 0300 GMT of the 9th; these four values had a difference from the observed of 1°. At 1500 GMT of the 9th the computed value was 66°; at 0300 GMT of the 10th, 68°; and at 1500 GMT of 10th, 69°; the observed values differed

[2] James' period was for only 4.5 years, thus a sufficient sampling of the population may not have been obtained.

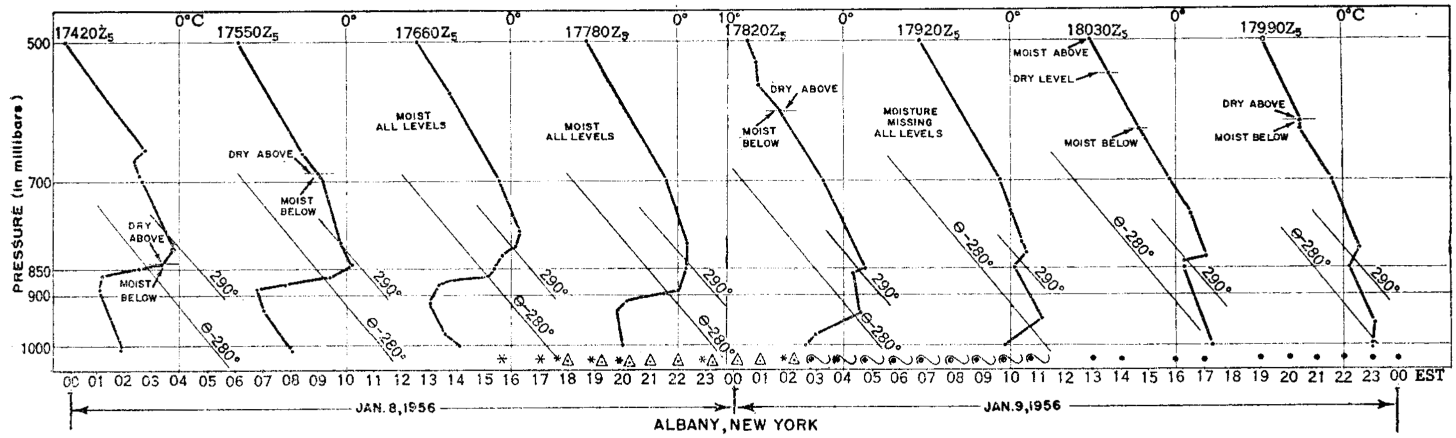


FIGURE 5.—The growth of the above-freezing area of the inversion layer is shown with the resulting change of types of hydrometeors at Albany, N. Y. for January 8–10, 1956. The pseudoadiabatic diagrams are plotted around the 0° isotherms spaced at 6 hour intervals with the hourly type of precipitation indicated beneath the soundings. Times are indicated in EST. 1000–500-mb. thickness values are at the top right of each sounding for comparison with type of precipitation.

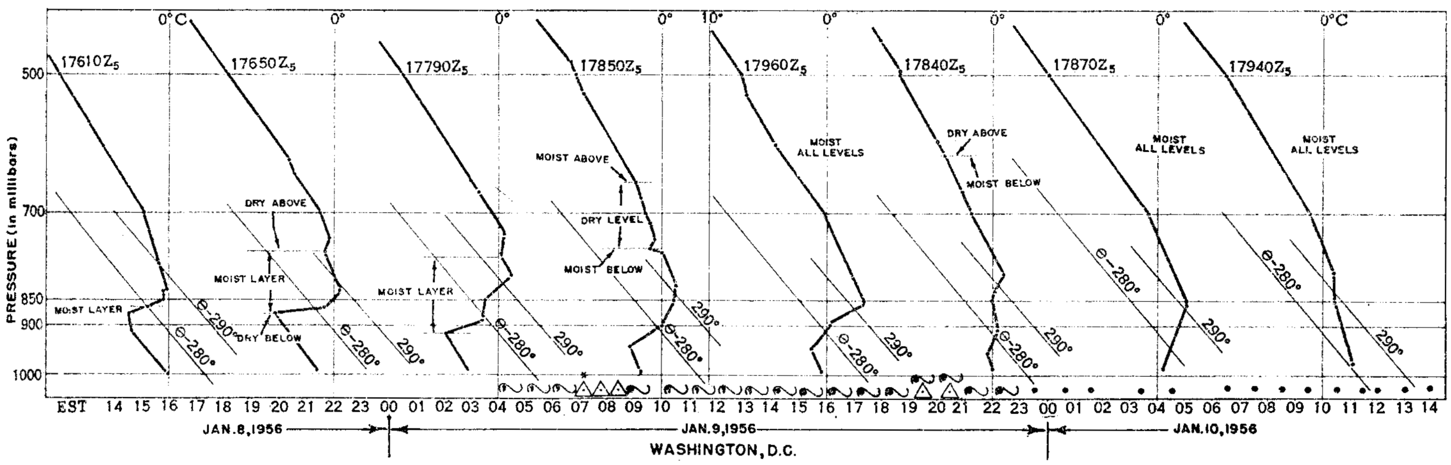


FIGURE 6.—Growth of the above-freezing area of the inversion layer with resulting change in type of hydrometeor, Washington, D. C. January 8–10, 1956.

by 2°, 1°, 0°, respectively. The formula for wave motion as derived by Rossby [9] and expanded by Cressman [10] was used in computing the stationary wavelength. The close agreement of the computed and observed values of the long-wave troughs indicated that a stable condition existed in the long-wave pattern upstream from the storm center during the period. This fact, plus the strong blocking effect over the northwestern Atlantic, explains why the storm center remained stationary off the eastern seaboard of the United States.

10. PSEUDOADIABATIC CHANGES

The vertical temperature distribution necessary for the formation of sleet or glaze is well known. Temperatures at the surface must be freezing or colder, coincident with an inversion above the surface that extends above the freezing point. Air above the inversion must have a high moisture content with wet bulb temperatures >0° C. so that evaporation of precipitation falling through this air cannot cool it below the freezing point. The initial form of precipitation, i. e., rain or snow, in the upper air matters but little, as the snow will melt in the above-freezing zone if the depth of this zone exceeds 2,000 to 3,000 feet, and warm air is continuing to be advected into the area. The height of the inversion above the surface generally determines the form of the hydrometeor reaching the surface: The greater the distance the more likely that the supercooled rain droplet will fall as sleet. The strength of the low-level winds is also important; strong winds and the resulting turbulence produce conditions more favorable for the formation of sleet. These conditions tend to maintain a deeper layer of cold air as well as creating a stirring action which increases the probability that the supercooled droplets will freeze.

Pseudoadiabatic charts for Albany, N. Y., and Washington, D. C. are shown (figs. 5 and 6, respectively) for each 6-hourly radiosonde, beginning when temperatures were sub-zero the entire height and ending with passage of the warm front and the change to a more normal type of hydrometeor. The soundings were traced directly from standard Weather Bureau pseudoadiabatic charts. Distances between the 0° C. isotherms are uniformly spaced with hourly intervals indicated. By this means it is possible to indicate the type-trend of the precipitation as the characteristics of the soundings change. Dew points were not plotted, but moist or dry air was indicated. In all cases of moist air identification the moisture content was saturated or within 4° C. of saturation.

The Albany sounding at 0400 EST of the 8th indicated a strong inversion layer near 850 mb., with moist air extending from the surface to slightly above the inversion layer, but with the entire sounding below freezing. During the next 12 hours, warm moist air advection had extended the near-saturation region over the entire sounding below 500 mb. and the upper portion of the inversion was above freezing. Snow flurries first were reported near 0800 EST but did not occur again until 1600 EST. The 1600 EST sounding showed a layer about 2,500 feet deep slightly above the zero isotherm beginning at the 800-mb. level. By 1800 EST sleet had become mixed with the snow, indicating a continued increase in the above-freezing layer and also a deep cold layer remaining near the surface. At 2200 EST the advection of warm moist air had produced a layer of above freezing temperature nearly 5,000 feet thick, and the precipitation was entirely sleet. Sleet changed to freezing rain at 0300 EST with the 0400 EST sounding indicating a below-freezing surface layer less than 2,000 feet deep. By 1000 EST only the immediate surface level was below freezing, and within the next 2 hours temperatures rose above 0° C. The warm front passed the station at that time, changing the freezing rain to occasional rain.

For Washington, D. C. (fig. 6), a similar picture can be observed, but with minor exceptions. The most important deviation from the Albany pattern occurred at Washington at 2200 EST of January 9. This sounding indicated a definite change from the preceding and following observations, with an almost isothermal lapse rate from the surface to near 8,000 feet. This change may have been caused by some brief deviation in the upper air flow which resulted in a decrease of the warm air advection over Washington, thus allowing the snow falling through this inversion layer to cool the air to near the 0° C. isotherm. Wexler, Reed, and Honig [11] have stated that cooling of the air by melting snow will first cause an isothermal layer to form aloft at the freezing point. This cooling aloft creates an unstable lapse rate just below the layer of melting snow, so that cold air may be transported downward. It might be well to point out that during this period of the near-isothermal lapse rate the precipitation on the surface was considerably heavier than it had been earlier, and that the types of hydrometeors varied from freezing rain to sleet to snow.

The 1000 EST sounding of the 9th indicated a small layer with a superadiabatic lapse rate. It is opined that this is the result of the temperature element acting as a wet bulb and therefore failing to register ambient temperatures correctly for a short distance upon entering a layer of dry air after passing through a saturated layer.

On these series of soundings from Albany and Washington, two inversions will be noted. In the early stages the potential temperature of these surfaces was near 300° and 288° A., respectively. The lower isentropic surface was associated with the warm front on the polar frontal system, and the upper potential temperature has been considered as the boundary of the tropical air aloft.

Various studies have been made of the correlation between the occurrence of snow or rain and the upper-air thickness values. It has been observed by Lamb [12] that a 1000–500-mb. thickness value of 17,300 feet is critical over the United Kingdom as the determining point between snow or rain-type precipitation. The cutoff point is not sharply defined but higher thickness values usually produce rain. Over the United States, the Analysis Center finds the critical thickness value for stations near sea level close to the same value of 17,300 over the northwestern United States where air-mass contrasts are weak, but nearer 17,800 feet thickness over eastern United States where strong air mass contrasts produce considerably more stability. However, the cutoff of the precipitation type again is not sharply defined and will range slightly over 100 feet either side of the 17,800-feet thickness height. The 1000–500-mb. thickness values are entered to the top right of each of the plotted soundings in figures 5 and 6.

11. STABILITY CONSIDERATIONS

Air flowing over the Northeastern States during this period was stable. This was established by examination and comparisons of the thicknesses of various layers of the soundings throughout the region, following the procedure developed by Showalter [13]. Thickness values for the following layers were considered: 1000–500 mb. (henceforth referred to as Z_5); 1000–700 mb. (Z_7), and 1000–850 mb. ($Z_{8.5}$). Figure 7A, B, and C, shows the result of the investigation, charting values of Z_7 minus $Z_5/2$ and values of $Z_{8.5}$ minus $Z_7/2$ for January 7 to 10. Under standard lapse rate conditions the value for $(Z_7—Z_5/2)$ is 550 feet and that for $(Z_{8.5}—Z_7/2)$ is 345 feet. Computed $(Z_7—Z_5/2)$ values of less than 500 feet indicate greater stability, while values greater than 500 feet indicate an increasing trend toward instability. Values near 300 feet are about the maximum of stability and near 700 feet the maximum of instability.

Similar values can be arrived at for $(Z_{8.5}—Z_7/2)$, but in this case, disregarding the sign, the reverse of the above is true. The higher values make for greater stability and the smaller values instability. The range in values here is from near 250 feet to about 550 feet, with the border line between the stable and unstable air near 370 feet. However, this latter value is variable, dependent upon the thickness of the $Z_{8.5}$ layer.

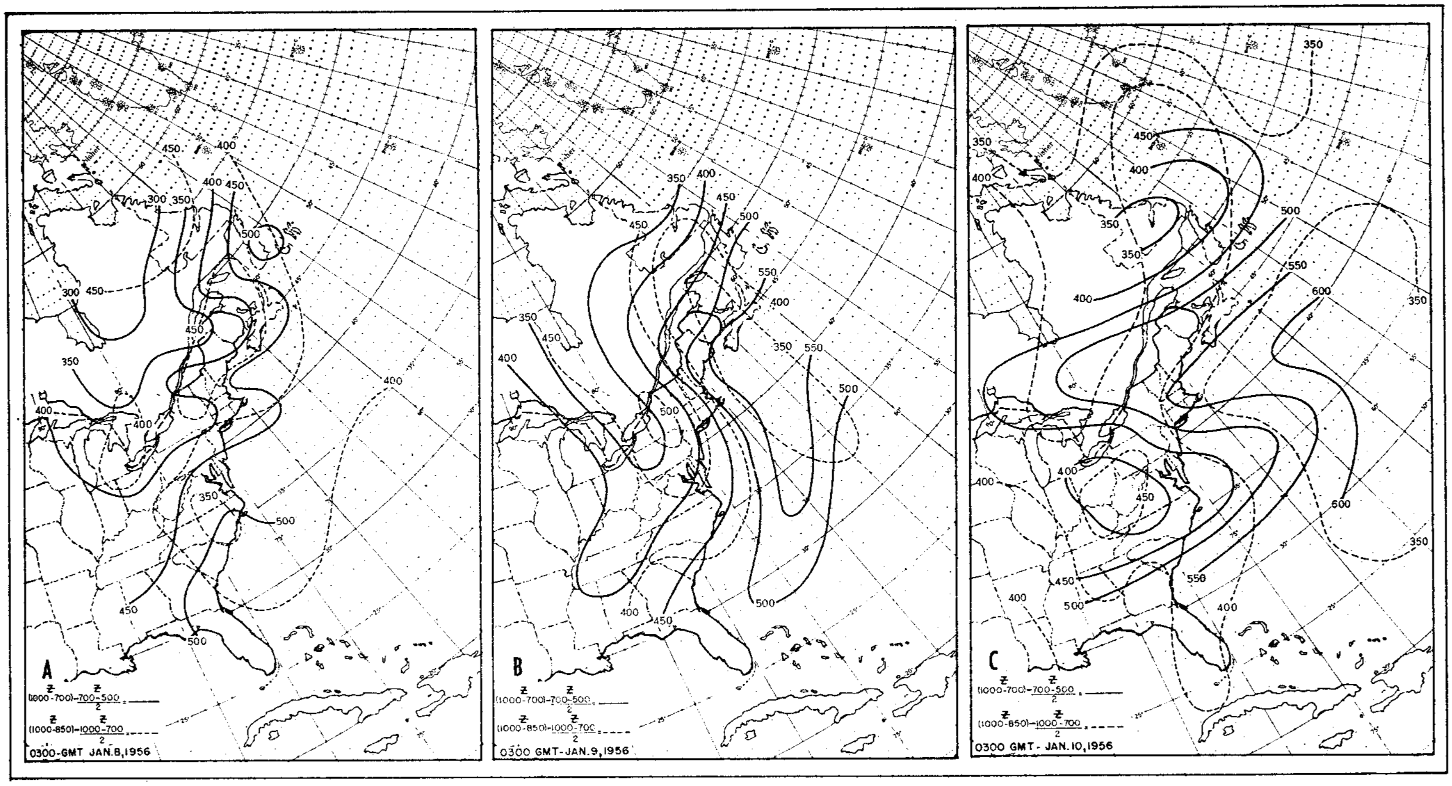


FIGURE 7.—Synoptic patterns of the stability and thickness relationships of two adjacent layers, January 8–10, 1956. (A) Composite chart for 0300 GMT, January 8. The 1000–700-mb. thickness value, minus the 1000–500-mb. thickness value divided by 2, is shown in tens of geopotential feet at 50-foot intervals by solid lines. Values below the 500-foot line indicate more stability and values above, a trend toward instability; the outside limits being near 300 and 700 feet, respectively. The dashed lines indicate the 1000–850-mb. thickness value, minus the 1000–700-mb. thickness value divided by 2, in tens of geopotential feet at intervals of 50 feet disregarding the sign, with departure from 370 feet indicating the reverse of the above, i. e., toward lower values would be increasing instability and higher values more stability. The limiting values for this layer are near 250 and 550 feet. (B) Composite chart for 0300 GMT, January 9, 1956. (C) Composite chart for 0300 GMT, January 10, 1956.

With these considerations in mind, it will be noted that the lines on the chart (fig. 7) indicate that the air from the surface to 500 mb. during this period was stable. However, there was some indication on the 10th that an area of instability was approaching the eastern seaboard in the vicinity of Boston. The value of the ($Z_{8.5}$—$Z_7/2$) in that area was on the border of becoming conditionally unstable; this was also true of the value at the higher level. This instability in the eastern quadrants of the Low was borne out by ships at sea which reported thunderstorms on the 10th and 11th.

12. PRECIPITATION

PRE-WARM FRONT

Precipitation totals from the sleet and freezing rain were for the most part small and generally under one-quarter of an inch. That such was the case may be easily understood in a résumé of what has been discussed previously. As indicated by the temperature range of the soundings, the moisture content was low, ranging from 4 to 6 grams per kilogram. This would yield in stable air and on a normal warm front slope between 0.25 and 0.50 of an inch per day [14]. Along the coastal sections the duration of the freezing precipitation was short, thus a large accumulation of ice was not obtained. Over the interior the total accumulation for the most part was also small, even though the period of occurrence was of greater duration than in the coastal sections.

POST-WARM FRONT

As shown previously by the upper air temperature comparisons and by the soundings at Albany and Washington, it appeared that a modified form of tropical air was over the northeastern seaboard States. The inversion associated with this air had a potential temperature near 296° or 300° A. at the beginning of the storm period, and gradually lowered to near the 292° or 294° potential temperature surface after the warm front passage. A cross section on the 9th at 1500 GMT (not shown) placed the 292° potential temperature surface at the following heights: Buffalo 7,000 feet, Albany 5,500 feet, Portland 5,000 feet, and Sable Island 2,500 feet. This would yield an approximate isentropic slope of 1 mile in 800 miles. This potential temperature surface could be extended to a possible weak surface warm front in the vicinity of 35° N., 58° W., oriented east-southeastward.

A slope of 1 to 800 with an easterly wind of 50–60 knots at the surface and 850-mb. level would yield an approximate vertical velocity of 0.03 meter per second. A vertical velocity of this value with dew points of 50°–55° F. should yield 24-hourly amounts of rainfall near 0.50 to 0.75 of an inch [14]. It is probable that with such a weak frontal slope this potential temperature surface was not uniform over the entire area. This in part would account for the variability of the precipitation totals. The coastal regions and adjacent areas received considerably larger rainfall totals than did the interior areas following the warm front passage and during the next several days.

Under conditions prevailing at that time, such a rainfall pattern would be considered likely. Strong onshore winds undoubtedly created considerable friction [15] along the coastal regions, while at the same time the air in that locality contained the greater amount of moisture.

Another indication for the continued heavier rainfall near the coast was the area of maximum anticyclonic vorticity displayed by thickness patterns over that region. (See fig. 1.) This is considered to be an area of maximum rainfall occurrence [16]. In this case it was not transported, but persisted over the coastal area from Boston to Portland throughout the period.

14. EFFECTS OF THE STORM

By far the most general effect of the ice storm was the delay and inconvenience inflicted on transportation facilities. The rash of minor traffic accidents at times exceeded one per minute in a few areas. Pedestrian accidents from falling on the slippery surfaces resulted in numerous broken bones, and bruises, as well as several deaths due to head injuries. Other deaths were ascribed to motor accidents. The glazing in Buffalo damaged power lines and trees. In several portions of New England the power companies had to produce heating on the high tension lines to thaw the accumulation of glaze. For short periods of 2 to 5 hours, it was necessary to discontinue bus service in certain communities, due to ice on roads and highways. Sanding and salting of the streets at times could not keep pace with the accumulation of glaze.

The long easterly fetch produced high tides from Delaware northward inflicting considerable damage on shore property and installations. At Atlantic City, N. J. tides of 4.5 feet above normal were recorded, and described as farther above normal than those of the 1954 and 1955 hurricanes. Philadelphia reported considerable damage from strong and gusty winds on the 10th. Blue Hill Observatory reported a peak gust of 65 m. p. h. on January 9, and a total rainfall that day of 3 inches. The arrival of the warm air and its persistence, in attendance with warm rains and melting of ice and snow, brought considerable flooding from ice-jammed streams.

The Harrisburg, Pa., Climatology Office of the Weather Bureau reports that the ice storm of the 8th and 9th was Statewide and estimates that the damage for the State was near $1 million. No heavy local losses were reported, but minor damage to roofs, trees, and automobiles was appreciable on a Statewide basis.

Dr. C. F. Brooks at Blue Hill Observatory, has informed us that at Mt. Washington, N. H., solid ice was built up to a maximum thickness of 6 feet on the northeast corner of the rampart but with an average thickness of 1 foot on the walls.

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